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The chronology of the Last Glacial Maximum and deglacial events in central Argentine Patagonia

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Abstract

This paper evaluates the chronology of the last glacial cycle and deglaciation in the Lago Pueyrredón valley of central Patagonia, 47.5º S, Argentina. The valley was a major outlet of the former Patagonian Ice Sheet and the moraines that record its fluctuations are an important proxy record of climate change in southern South America. Such moraines are well-preserved in the Lago Pueyrredón valley owing in part to the semi-arid environment east of the mountain front. Here, we provide the first direct chronology for the age of the "Rio Blanco" moraine system by utilizing cosmogenic-nuclide surface exposure ages. Boulders on the moraines give ^{10}Be exposure ages that indicate the Last Glacial Maximum (LGM) maximum extent occurred by 27 – 25 ka. Subsequent advances occurred at ca. 23 ka, ca. 19 ka, and ca. 18 – 17 ka. Initial deglaciation began after 18 – 17 ka and was interrupted as evidenced by the Lago Columna moraines upvalley. Subsequently the outlet glaciers occupying both the Lago Pueyrredón basin (Chilean name: Lago Cochrane) and the Lago Buenos Aires basin (Chilean name: Lago General Carrera) to the north, rapidly retreated more than 80 kilometers at around 16.5 – 15 ka. The timing of the LGM maximum extent and the onset of deglaciation occurred broadly synchronously throughout Patagonia. Deglaciation resulted in a series of interconnected glacier-dammed lakes in the region that initially drained toward the Atlantic Ocean and later drained to the Pacific Ocean as a consequence of disintegrating ice in the Andes.

Keywords: Cosmogenic nuclide surface exposure dating; Last Glacial Maximum; Glacial chronology; southern South America; Beryllium-10; Patagonia; Antarctic Cold Reversal.

1. Introduction

The purpose of this paper is to establish the chronology of the last glacial and deglacial events in the Lago Pueyrredón valley, 47.5º S, Argentina (Fig. 1). Specifically, we use cosmogenic ^{10}Be surface exposure ages to determine the age of the "Río Blanco" moraine system [Caldenius, 1932], which is the innermost and youngest of a series of moraine groups preserved on the eastern margin of this former lobe of the Patagonian Ice Sheet. Prior to this study [and Hein *et al*., 2009], no direct quantitative age data existed for these moraines, except for limiting ^{14}C data [Wenzens, 2005]. Overall, few studies have been carried out in the valley since the time of Caldenius [1932] with notable exceptions ([Feruglio, 1950], [Glasser and Jansson, 2005] and [Wenzens, 2005]). With detailed mapping and a new chronology, and by drawing on previously published data, we reconstruct the timing and extent of outlet glaciers in the Lago Pueyrredón and nearby valleys during and following the LGM, and link the process of deglaciation with the regional paleolake record, which has implications for continental-scale drainage diversion as the ice disintegrated in the Andes.

The southern Andes span a wide range of latitudes and associated climate systems in southern South America. Located between 37º S and 56º S, the Patagonian and Fuegian Andes form a significant barrier to the precipitation-bearing southern westerly winds and to Southern Ocean circulation [Coronato *et al.,* 2008]. They now host three large icefields, the Hielo Patagónico Norte (North Patagonian Icefield), the Hielo Patagónico Sur (South Patagonian Icefield) and the smaller Darwin Cordillera Icefield, which are remnants of a much larger Late Pleistocene mountain ice sheet that covered the southern Andes [Rabassa, 2008]. The moraines that mark the former extents of Patagonian glaciers are an important proxy for past terrestrial climates. Such records, if well-dated, can give insight into past climate changes [e.g., Denton *et al.,* 1999a]. When compared to the millennial-scale climate changes observed in marine cores (e.g., Heinrich events) and ice cores (e.g., Dansgaard-Oeschger cycles), they can give insight into the processes and mechanisms that drive regional and global climate changes (e.g., [Denton *et al.,* 1999a] and [Kaplan *et al.*, 2004]). The behavior of glaciers may be also in part dependent on glacier-flow dynamics unrelated to climate, such as surging, tidewater calving-margin dynamics, and the influence of topography (e.g., [Glasser and Jansson, 2005], [Rivera *et al.*, 1997], and [Warren and Sugden, 1993]). Thus, not all glaciers respond linearly to a climate change. In addition, moraines that are preserved may vary between valleys due to the extent of subsequent glacial advances and geomorphic processes. Thus, where possible, developing multiple glacial chronologies in close proximity helps to ascertain the overriding climate signal from more local topographic and other influences.

The climate in central Patagonia during and following the LGM is more uncertain than the climate in northern and southern Patagonia [Rabassa, 2008]. Part of the reason for the relative lack of glacier-based climate reconstructions during the LGM is that glaciers often terminated in the sea on the west coast and therefore the moraine record is limited. In contrast, the eastern outlets terminated on land and are well-preserved, but the arid environment supports little organic material for radiocarbon age dating. The environment is, however, suitable for cosmogenic-nuclide studies, and the most detailed LGM glacial chronology in central Patagonia utilizes surface exposure ages from boulders on moraines located east of Lago Buenos Aires, 46.5º S, Argentina (Fig. 1). Here, the Fenix V – I moraines have surface exposure ages (^{10}Be,

^{26}Al) ranging from ca. 23 – 16 ka ([Douglass *et al.*, 2006] and [Kaplan *et al.*, 2004]). The youngest Menucos moraine, which is also located at the eastern end of Lago Buenos Aires, but is separated from the Fenix moraines by lake deposits, is dated at 14.4 ± 0.9 ka based on ^{10}Be exposure ages [Douglass *et al.*, 2006].

While the cosmogenic-nuclide data indicate that the youngest moraines at Lago Buenos Aires mark glacier advances during the last major glacial period, Wenzens [2006] argues instead that the moraines were deposited during the Late-glacial to Holocene period. This alternative interpretation is based on associations made with the timing of glacial advances in outlet valleys further south, and on two limiting radiocarbon ages (8.6 ^{14}C ka and 5.9 ^{14}C ka) from high paleoshoreline terraces within the limit of the Menucos moraines [Wenzens, 2006]. The fact that such a difference of opinion can exist over the timing of major glacial advances reflects in part the need for additional data in the region. Thus, to define better the central Patagonian glacier response during and following the LGM and develop regional glacial-climate signals, we focus on the chronology for moraines in the Lago Pueyrredón valley. In addition to providing insight into the glacial history of the region, these data will provide constraints on high-resolution glaciological models (e.g., [Hubbard, 1997], [Hubbard *et al.*, 2005] and [Hulton *et al.*, 2002]) that can test the time-transgressive properties of the ice sheet and its link to climate as a whole [Hulton *et al.*, 2003].

2. Regional setting

The central Patagonian region refers to the area surrounding the North Patagonian Icefield, roughly between latitudes 45º S and 48º S. The present icefield is the second largest temperate icefield in the Southern Hemisphere with an area of ca. 3,950 km^2

[Barcaza *et al.*, 2009]. It is sustained by abundant westerly precipitation (4 – 10 m a^{-1}) which dissipates rapidly in the lee of the Andes to semi-arid levels (ca. 200 mm a^{-1}) within a few tens of kilometers of the mountain front ([Inoue *et al.,* 1987] and [Prohaska, 1976]). The region surrounding the North Patagonian Icefield is characterized by a series of glacially deepened valleys and fjords whose orientation is largely controlled by underlying geologic structures [Glasser and Ghiglione, 2009]. East of the North Patagonian Icefield are two large, broadly west-east trending valleys that have been over-deepened through glacial erosion (Fig. 1) [Rabassa, 2008]. The cross-border lakes of Lago Buenos Aires (Chilean name: Lago General Carrera) and Lago Pueyrredón (Chilean name: Lago Cochrane) occupy these deep depressions, with lake depths of ca. 600 m and > 400 m, respectively [Murdie *et al.,* 1999]. Geomorphic mapping of scoured bedrock and glacial lineations in the Lago Cochrane/Pueyrredón and neighboring valleys indicate that they were occupied by topographically controlled, fast-flowing outlet glaciers [Glasser and Jansson, 2005]. Thus, the valleys were major outlets of former, expanded Patagonian Ice Sheets.

The Lago Pueyrredón valley, the main focus of this study, is bounded by basaltic upland areas of the Meseta del Lago Buenos Aires to the north, and the Mesetas Belgrano and Olnie to the south. The surrounding geology consists of Jurassic silicic volcanics and Cretaceous to Miocene mixed marine and continental sediments [Gorring *et al.*, 2003]. To the west, these overly highly deformed Paleozoic basement rocks of the Eastern Andes Metamorphic Complex, which are intruded and deformed by the Mesozoic Patagonian batholith and later plutonics in the Andean Cordillera ([Lagabrielle *et al.*, 2004], [Ramos and Kay, 1992] and [Suárez and De La Cruz, 2001]). Quaternary glacial and glaciofluvial sediments fill the basin 150 – 200 kilometers east of present day outlets glaciers of the North Patagonian Icefield [Caldenius, 1932].

2.1. Existing glacial chronology

Following Caldenius [1932], the moraines east of Lago Pueyrredón are grouped into four major moraine systems named in decreasing relative age and eastward extent, the "Gorra de Poivre", "Cañadón de Caracoles", "Hatcher", and "Río Blanco" (Fig. 2). The oldest moraine system is associated with the "Greatest Patagonian Glaciation" which occurred at ca. 1.1 Ma ([Coronato *et al.*, 2004], [Mercer, 1976], [Singer *et al.*, 2004] and [Ton-That *et al.*, 1999]). Until recently, the only age constraint for the moraine systems came from magnetic polarity measurements in glacial sediments [Sylwan *et al.*, 1991] which indicated that part of the Cañadón de Caracoles moraine system was deposited during the reversed Matuyama chron at more than 780 ka ([Shackleton *et al.*, 1990] and [Singer and Pringle, 1996]). Recently, the age of the penultimate Hatcher moraine system was determined by cosmogenic-nuclide surfaces exposure methods at ca. 260 ka or Marine Isotope Stage 8 [Hein *et al*., 2009].

Initial surface exposure ages of 27 – 25 ka from boulders on the outermost of the Río Blanco moraines support deposition during the LGM [Hein *et al*., 2009]. However, previous radiocarbon ages have been interpreted to suggest that the Río Blanco moraines instead represent a series of Late-glacial to Holocene glacier advances ([Wenzens, 2002], [Wenzens, 2005] and [Wenzens, 2006]). The basis for this interpretation rests on three key radiocarbon ages (Fig. 2). The first age is from a mollusk shell located in lake sediments at the foot of the Cañadón de Caracoles escarpment; this yielded an age of 17.2 ^{14}C ka, suggesting the Hatcher moraines that impounded the lake represent the LGM. The second age of 11.8 ^{14}C ka [Mercer, 1982] from peat in the Cañadón de Caracoles drainage channel provides a minimum age for the Río Blanco moraines whose meltwater drained via this route. A third age from the humic base of a kettle hole associated with the younger Río Blanco moraines gives a minimum limiting age of 8.3 ^{14}C ka BP. Wenzens' interpretation of the radiocarbon ages conflicts with data indicating that ice was limited to the cordillera during the Holocene.

Evidence for relatively restricted Holocene ice in the region was first established by Mercer [1976] who obtained a radiocarbon age of 11.0 ^{14}C ka from near the present day terminus of Glacier Témpano, a northwestern outlet of the South Patagonian Icefield. Further evidence in support of relatively restricted Holocene ice east of the North Patagonian Icefield is found in the Nef valley (Fig. 1b) where a shell and gyttja within a kame delta have calibrated radiocarbon ages of 13.8 – 13.5 ka and 13.0 – 12.7 ka, respectively, and boulders on the delta have surface exposure ages of 10.5 ± 1.3 ka and 12.2 ± 1.5 ka [Turner *et al.*, 2005]. In the Río Bayo valley, an OSL age of 9.7 ± 1.0 ka from a stratum in a kame deposit, together with two surface exposure ages from boulders in the valley bottom, with a mean of 10.9 ± 1.0 ka, indicate an advance of ca. 20 km from present day ice margins during the Late-glacial or early Holocene period [Glasser *et al.*, 2006]. Finally, boulders on the Fachinal moraines, which were deposited on an active delta now raised above Lago General Carrera, have weighted mean surface exposure ages of 6.2 ± 0.8 ka (n=6; excluding one outlier) and 8.5 ± 0.7 ka (n=9; excluding three outliers) [Douglass *et al.*, 2005]. Late Holocene and Neoglacial advances have been identified on both sides of the North Patagonian Icefield based mainly on radiocarbon ages and dendrochronology (e.g., [Aniya, 1995], [Glasser *et al.*, 2002], [Harrison *et al.*, 2008], [Mercer, 1982] and [Winchester *et al.*, 2001]). These deposits are generally situated within about fifteen kilometers of present-day ice margins [Glasser *et al.,* 2005].

2.2. Paleolakes

A 100 km kilometer break in the chain of the Andes at ca. 48º S separates the present day North and South Patagonian Icefields, exposing a low-level drainage route to the Pacific Ocean from the east side of the Andes. The Río Baker occupies this depression and drains the eastern outlets of the North Patagonian Icefield, Lago Buenos Aires/General Carrera and Lago Pueyrredón/Cochrane (Fig. 1b). The present drainage divide is located near the Fenix and Río Blanco moraine systems at the eastern ends of Lago Buenos Aires and Lago Pueyrredón, respectively, about 50 kilometers east of the mountain front [Caldenius, 1932]. Therefore, despite their position on the eastern side of the Andes, the lakes drain to the Pacific Ocean via the Río Baker. During glacial times when mountain glaciers expanded and coalesced into a continuous ice sheet, this low-level drainage route was blocked, forcing a 200 km shift in the drainage divide toward the cordillera, and causing meltwater from the Lago Buenos Aires and Lago Pueyrredón outlet glaciers to drain to the Atlantic Ocean ([Bell, 2008], [Mercer, 1976] and [Turner *et al.,* 2005]) (Fig. 1a). As a consequence, a series of interconnected pro-glacial lakes formed in response to the disintegration of ice in the Andes following the LGM. Thus, the formation of paleolakes and their chronology provides important insight into deglaciation and continental-scale lake diversions in the region.

The geologic evidence indicates the presence of paleolakes at heights ranging from about 100 meters to about 500 meters above the present-day elevation of Lago Buenos

Aires (~200 m) and Lago Pueyrredón (~150 m) ([Bell, 2008] and [Turner *et al.*, 2005]). The evidence includes paleoshorelines, beaches, terraces, raised deltas, and lake sediments ([Bell, 2008], [Caldenius, 1932], [Douglass *et al.*, 2005], [Turner *et al.*, 2005] and [Wenzens, 2005]). The former lake levels were investigated across a broad area by Turner *et al.* [2005], who recorded the elevation of a number of terraces in the region surrounding the town of Cochrane, Lago Cochrane and Lago General Carrera. They distinguish two "united" paleolake-levels throughout the region, and a third, higher paleolake level restricted to the Lago Cochrane/Pueyrredón basin. The elevation of the upper paleolake ranges between 654 – 633 meters. The elevation of the upper *united*-paleolake ranges between 512 – 489 meters and between 484 – 390 meters in the Lago Cochrane and Lago General Carrera basins, respectively. The elevation of the lower *united*-paleolake ranges between 397 – 375 meters and between 355 – 305 meters in the Lago Cochrane and Lago General Carrera basins, respectively. The range in paleolake elevations within and between the two basins is attributed to different rates of isostatic recovery and tilting. Assuming the highest shoreline remnants represent a single lake level, the upper-paleolake elevation decreases towards the east at ~0.6 m/km. The mean elevation of the lower united-paleolake remnants decrease eastwards at ~0.4 m/km [Turner *et al.*, 2005].

Turner *et al.* [2005] sampled peat and plant macrofossils from a range of ice-contact and associated deposits to establish a lake chronology. Cores were taken from two kettle holes associated with the lower united-paleolake at ca. 370 meters elevation near the base of Cerro Ataud, about 110 kilometers west of the Río Blanco moraine system (Fig. 1b). The peat samples in each core yield calibrated radiocarbon ages of 16.3 ± 0.2 ka and 14.8 ± 0.9 ka, and the plant macrofossils from the same cores give ages of 15.7 ± 0.3 ka and 13.6 ± 0.2 ka, respectively. Turner *et al.* [2005] interpret the data as evidence for early deglaciation near Cerro Ataud. Given that the united-paleolake could only exist when ice had withdrawn over 100 kilometers to expose the valley linking the two lake basins (Fig. 1b), and that the youngest Fenix moraine is dated at ca. 16 ka [Kaplan *et al.*, 2004], Turner *et al.* [2005] suggest a rapid retreat of the outlet glaciers occurred at about 16 – 15 ka. Further, they suggest that the lower united-paleolake drained to the Pacific Ocean at ca. 12.8 cal ka during the Younger Dryas, an age based on the youngest direct radiocarbon age on lake shore sediments.

A second focused study on the paleolake-levels was conducted by Bell [2008]. Bell [2008] identifies a series of seven raised lacustrine braid deltas at elevations ranging from 100 meters to 450 meters above Lago General Carrera. Bell argues that the deltas, with steeply inclined subaqueous foreset beds and nearly flat topset beds, were formed by the punctuated drainage of the lake, possibly related to the breaking of an ice dam causing the paleolake to suddenly drop by 30 to 150 meters, which was followed by periods of stability. Bell identifies a series of small terraces (1-2 m wide) on the exposed delta fronts which he interprets to represent wave erosion during the falling lake levels. A two meter thick layer of volcanic ash on Bell's "Delta 2", which is equivalent to the lower united-paleolake of Turner *et al.* [2005], represents the 6.7 ka BP eruption of Volcan Hudson ([Bell, 2008] and [Naranjo and Stern, 1998]). This ash layer is not found on the present day delta ("Delta 1"). Thus, Bell suggests that the lower united-paleolake was still in existence at this time in the Holocene.

This is in agreement with the age of moraines on the raised delta at Fachinal [Douglass *et al.,* 2005], which is at an elevation equivalent to the lower united-paleolake, but it too conflicts with the radiocarbon ages presented by Turner *et al.* [2005] which instead suggest the lake had drained by ca. 12.8 cal ka. The recent age determination for the Menucos moraines at 14.4 ± 0.9 ka also conflicts with some of the radiocarbon ages of macrofossils ([Douglass *et al.*, 2006] and [Turner *et al.*, 2005]). Thus, the age of the lakes is also controversial.

3. Materials and methods

3.1. Geomorphic mapping

The geomorphic map of the Lago Pueyrredón valley in Figure 2 is based on interpretation of satellite imagery and digital elevation models, backed up by field observations made during three field visits between 2006 and 2009. The datasets are listed in Table 1. The digital elevation model was artificially illuminated using ESRI's ArcGIS software in order to accentuate relief such as scarps and moraine ridges [cf. Ganas *et al.*, 2005]. The elevations presented are based on GPS and digital barometric altimeter measurements made in the field (assumed uncertainty ± 10 m) and derived from the Shuttle Radar Topography Mission (90m) digital elevation model [vertical error < 16 m; Farr *et al.*, 2007] referenced to modern sea level. Our mapping is similar to previous work by Wenzens [2005], and reveals a series of moraines, outwash terraces, paleoshorelines and landslide deposits. The following sections describe the moraines and paleoshorelines and clarify some interpretations necessary for sampling.

3.1.1. Río Blanco and recessional moraines

A series of moraines were mapped and distinguished by their surface morphology and sedimentological characteristics. The moraines, with arcuate plans and regular gradients, vary from hummocky continuous landforms with slope angles of about 17º - 20º and relief of 20 – 50 meters (Fig. 3a), to distinct, sharp-crested, continuous landforms with slopes of 20º - 28º and relief of 20 – 40 meters (Fig. 3b). Boulder weathering ranges in severity from minor aeolian erosion and spalling (depth < 2 cm), to surfaces that are smoothly polished. Our mapping of the Río Blanco moraine system reveals three distinct glacier advances, here termed "ice limits". Each ice limit is composed of several individual moraines that reflect readvances or standstills. The moraines are best preserved in the southern half of the valley. In the north, meltwater sourced from the Lago Ghio lobe removed most of the moraines associated with the outermost limits [cf. Wenzens, 2005]. The First Limit marks the maximum extent of the Río Blanco moraines. The Second Limit moraines cross-cut these in places producing locally high relief of 20 – 30 meters. The Third Limit is continuous and encompasses the main basin.

On the south flank of the valley, a bedrock step of 150 – 200 meters isolates a basin from the main valley (Fig. 2 and 4); the basin extends 8 kilometers eastward. A series of arcuate terminal moraines of the Second Río Blanco limit enclose the basin to the east and are mantled by lacustrine sediments below 625 meters. A sediment section at 570 meters elevation reveals till overlain by a two-meter thick sequence of deformed, varved lacustrine sediments containing dropstones (Fig. 3c); these are overlain by about one meter of till which is in turn covered by about 50 cm of fine sediments. These data indicate the formation of a pro-glacial lake between the First and Second and Third readvance limits.

To the west of the Río Blanco moraines some 40 kilometers up valley are younger probable ice limits. A distinct causeway 300 meters wide by four kilometers long separates Lago Posadas from Lago Pueyrredón (Fig. 2). It is composed of well-rounded cobbles and pebbles and is covered by sand dunes to the south. The density of erratic boulders situated on the surrounding volcanic bedrock is higher in close proximity to the causeway (Fig. 3d). The causeway has the planimetric form of an asymmetric cross-valley moraine [Andrews and Smithson, 1966] and we interpret the feature as a moraine that marks the grounding line of a glacier that terminated in a pro-glacial lake. The eastern shore of Lago Posadas down-valley displays similar morphology and could also reflect a former ice limit. These moraines are correlated to those in the north near Lago Columna, which were deposited by the glacier occupying the Chacabuco valley.

3.1.2. Paleoshorelines

In satellite imagery and in the field, paleoshorelines are commonly marked by a distinct 5 – 10 meter scarp at a constant elevation; these encompass the eastern end of the basin inside the Río Blanco moraine limits (Fig. 5). Multiple closely-spaced beaches with well-rounded cobbles parallel the scarps at some locations. A series of deltas grade to the shorelines, and wave-polished boulders and rock shelters occur on the lowest shoreline at Cerro de los Indios (Fig. 6). Three distinct shoreline levels are distinguishable at 505 – 470 meters, 400 – 370 meters and 300 – 270 meters. The spillway for the Cañadón de Caracoles, which is the only drainage route in the Lago Pueyrredón valley toward the east, is situated at ~475 meters. The two lower shorelines can be identified in the adjacent Lago Buenos Aires valley at elevations of 400 – 370 meters and 300 – 270 meters (Fig. 5b). The spillway for the Río Deseado, which is the only drainage route in the Lago Buenos Aires valley toward the east, is situated at ~380 meters.

Our interpretation is that the upper paleolake coincides with a time when a glacier occupied the Lago Pueyrredón valley and drainage of the pro-glacial lake was directly to the Atlantic Ocean via the Cañadón de Caracoles. The lower two paleoshorelines represent a “united lake” that formed when glaciers occupying both valleys had withdrawn sufficiently to allow Lago Pueyrredón to drain into Lago Buenos Aires. The upper united-paleolake drained to the Atlantic Ocean via the Río Deseado at Lago Buenos Aires, while the lower united-paleolake drained westward to the Pacific Ocean [Bell, 2008]. If these united-paleolake levels are analogous to the higher elevations identified by Turner *et al.* [2005] further to the west, then the similarity in elevations indicate that isostatic recovery is both less than in the west and uniform at the eastern ends of the lake basins. The average tilt in the paleoshoreline levels, which was reported on a line ENE along the axis of Lago General Carrera by Turner *et al.* [2005], fits the shorelines on the eastern end of Lago Buenos Aires, but the tilt is steeper for the Lago Pueyrredón basin. The range of elevations at each paleolake level of ca. 30 m could represent its gradual lowering, perhaps resulting from incision of the drainage channel causing a lowering of the spillway. In the following sections, we make the basic assumption that the paleolake levels represent a single lake that lowers as a response to progressive stages of deglaciation.

3.2. Surface exposure methods

Moraine boulders (mainly granites) were sampled to obtain ^{10}Be exposure ages for the Río Blanco ice limits (Fig. 7). The nearest granitic rocks are within the San Lorenzo Plutonic Complex, about 80 kilometers from the innermost moraines [Suárez and De La Cruz, 2001]. Samples were collected with hammer and chisel following standard sampling protocols [Gosse and Phillips, 2001]. For example, preference was given to large, flat topped, and stable boulders located on moraine crests, and the top few centimeters were sampled away from edges and pitting, spalling or ventifaction. Shielding by local topography was measured using an inclinometer. Vegetation is limited to small shrubs, and historical snow accumulation has been thin and short-lived in this semi-arid [Prohaska, 1976] and windy environment. Moreover, models predict increased aridity during glacial times [Hulton *et al.*, 2002]. Therefore, given the large (> 75 cm) and exposed moraine boulders sampled in this study, we consider post-depositional shielding of the boulder surface by snow or vegetation to be negligible.

Moraine boulders marking each distinct ice limit were sampled at the locations illustrated in Figure 2. The samples were crushed and sieved to obtain the 250 – 710 μm size fractions for analysis. The extraction and preparation of BeO targets was conducted at the University of Edinburgh's Cosmogenic Isotope Laboratory following procedures adapted from the methods of Bierman *et al.* [2002], Kohl and Nishiizumi [1992] and Ivy-Ochs [1996] [see Hein, 2009]. The AMS analyses were performed at the Scottish Universities Environmental Research Centre (SUERC). All measurements are standardized relative to 07KNSD [Nishiizumi *et al.*, 2007]. The ^{10}Be exposure ages (Table 2) were calculated using the CRONUS-Earth web-based exposure age calculator (version 2.2; see [Balco *et al.*, 2008] for full documentation)[1] which implements the revised ^{10}Be standardization and half-life (1.36 Ma) of Nishiizumi *et al.* [2007]. The revised half-life has little effect (~1%) on exposure ages for the LGM. Sample thickness (Table 3) and density (2.7 g cm^{3}) are used by the exposure age calculator to normalize ^{10}Be concentrations to the boulder surface. The

[1] http://hess.ess.washington.edu/math/

exposure age calculator's "standard atmosphere" default setting is used for the elevation – pressure relationship; this incorporates geographically variable mean sea-level pressure and 1000 mbar temperature fields obtained from the NCEP-NCAR reanalysis [Balco *et al.,* 2008].

To facilitate cross-comparison with other cosmogenic-nuclide chronologies, we report exposure ages based on all scaling models implemented in the exposure age calculator ([Desilets *et al.,* 2006], [Dunai, 2001], [Lal, 1991]/[Stone, 2000], and [Lifton *et al.,* 2005]). The range in exposure ages derived using the different scaling models is less than 6% (Table 2), which is less than the current uncertainty in the ^{10}Be production rate value itself (> 10%; [Balco *et al.,* 2008], [Balco *et al.,* 2009], and [Putnam, submitted]). The exposure ages discussed in this paper are calculated according to the Dunai [2001] scaling model with a reference (sea level high latitude) ^{10}Be spallogenic production rate of 4.43 ± 0.52 atoms $g^{-1}(Si0_2)$ a^{-1} [±1σ; Balco *et al.,* 2008] with muonic production of ^{10}Be calculated according to Heisinger *et al.* [2002a, 2002b]. Unless stated otherwise, 1σ analytical uncertainties are reported. For means, the uncertainty is based on an average of the upper and lower age limits. The upper limit is the exposure age plus one sigma, and the lower limit is the exposure age minus one sigma. No correction is applied for post-depositional shielding or for rock surface erosion, except where noted; however, the effect of including the long-term maximum erosion rate estimated at nearby Lago Buenos Aires [1.4 mm ka^{-1} for surfaces older than 760 ka; Kaplan *et al.*, 2005], would be to increase exposure ages by about 2% and 3% for the youngest and oldest exposure ages reported, respectively.

Given the proximity to and associations with glaciation in the Lago Buenos Aires valley, in Table 4 we recalculate some of the ^{10}Be surface exposure ages published by

Kaplan *et al.* [2004] and Douglass *et al.* [2006]. The recalculation makes a direct comparison of the data sets possible because the samples were taken from a similar latitude, elevation and geomorphic environment and because they are based on the same production rate, scaling factor and underlying geologic and geomorphic assumptions. Thus, systematic uncertainties for scaling factors are reduced. Most important, any changes in the production rate value itself will not affect comparisons made between the data sets. Data from the Strait of Magellan originally published by McCulloch *et al.* [2005] and Kaplan *et al.* [2008] are recalculated in the same manner (Table 4). Radiocarbon ages presented from this point forward are reported as calibrated years before 1950 using the calibration curve of Fairbanks *et al.* [2005; v. Fairbanks0107].

4. Results

The cosmogenic-nuclide results are listed in Tables 2 & 3 and are illustrated in relation to the moraines in Figure 8. The samples marking the First Limit range from 32.2 – 25.4 ka. The oldest sample (BC07-8; Fig. 7a) falls outside of 2σ uncertainty of the remaining population. This sample may contain inherited nuclides from a previous exposure or it marks an earlier advance that terminated at the same location. Excluding this outlier from the remaining population reduces the range to 27.1 – 25.3 ka, with an arithmetic mean age of 26.0 ± 0.9 ka. The outermost sample marking the Second Limit yields an age of 22.9 ± 0.7 ka. The two innermost samples yield ages of 22.3 ± 0.8 ka and 23.1 ± 1.3 ka. The indistinguishable ages obtained from boulders on the outer and innermost moraines of the Second Limit suggest that the small lake that would have shielded the innermost samples from cosmic radiation by about 65 meters of water, must have been relatively short-lived (Fig. 4). The arithmetic mean age is 22.8 ± 1.0 ka. The samples of the Third Limit yield ages of 19.9 – 13.2 ka. Sample BC07-18 (13.2 ka) is a clear outlier (3σ). This sample was perched near the edge of a steep-sided moraine (28º) and may have shifted after deposition. Excluding this sample, the range is 19.9 – 17.8 ka with an arithmetic mean age of 19.1 ± 0.7 ka. The final sample from the youngest Río Blanco moraine yields an age of 17.4 ± 0.7 ka. Given the highest lake shoreline is immediately adjacent to this moraine, we assume that this is the final glacial advance prior to deglaciation. Finally, sample BC07-13 located on bedrock below the paleolake levels yields an age of 15.9 ± 0.8 ka.

The surface exposure ages are internally consistent and concur with the relative ages of the moraines. The oldest ages come from moraine boulders situated on the outermost and oldest moraine limit, and the youngest ages come from boulders on the innermost and youngest moraine limits (Fig. 8). Given their internal consistency, we consider the ^{10}Be exposure ages representative of the approximate timing of each glacial advance. The single exposure ages that date the final advance (BC07-21) and deglaciation (BC07-13) are broad approximations given the potential complications that are inherent with exposure ages of a single sample. Nevertheless, the internal consistency gives a measure of confidence in their ages.

5. Discussion

5.1. Río Blanco moraines – revised chronology

The surface exposure ages suggest that the moraines were deposited over a 7 – 10 ka period with key advances at ca. 26 ka, ca. 23 ka and ca. 19 ka, and with deglaciation beginning after ca. 18 – 17 ka. The data indicate that all of the Río Blanco moraines were deposited during the Last glacial period and not during the Late-glacial to Holocene as previously reported [Wenzens, 2005]. The argument for a Holocene advance is partly based on a minimum limiting radiocarbon age of 9.3 cal ka (8.3 ^{14}C ka; Fig. 2). However, if a glacier occupies the Lago Pueyrredón valley, meltwater drains eastward via the Cañadón de Caracoles (Section 3.1.2), and radiocarbon ages indicate this outlet had dried before ca. 13.7 cal ka (11.8 ^{14}C ka; [Mercer, 1976], [Mercer, 1982] and [Turner *et al.,* 2005]). This is inconsistent with a Holocene advance to the eastern end of Lago Pueyrredón. Moreover, in this arid environment, organic material may not begin to accumulate immediately following moraine deposition (or channel drying), but instead may reflect the onset of more humid conditions ([McCulloch *et al.*, 2005] and [Turner *et al.*, 2005]). Therefore, the ^{14}C ages based on the accumulation of organic material may not necessarily be close minimum ages for deglaciation in this semi-arid environment on the eastern side of the Andes.

The glacier occupying the Lago Pueyrredón basin retreated more than 80 kilometers at ca. 16 ka, an age based on an exposure age from an erratic boulder (BC07-13: 15.9 ± 0.8 ka) resting on bedrock at an elevation of ca. 350m (Fig. 7d). This sample has been submerged in a pro-glacial lake and thus it would have been shielded from cosmic radiation by 100 – 150 meters of water, reducing spallation and muonogenic production of ^{10}Be by over 99.9%. Subsequently, it was exposed after the lake drained into Lago Buenos Aires. Thus, we argue the exposure age of this sample is best interpreted as a minimum age for the draining of Lago Pueyrredón, and by inference, for the drying of the Cañadón de Caracoles drainage channel.

5.2. Regional significance of the new chronology

We date the maximum extent of ice during the last glacial period in the Lago Pueyrredón valley by 27 – 25 ka; the mean of three samples is 26.0 ± 0.9 ka. This mean age excludes the one ~32 ka sample, which interestingly was found on a small crest just exterior to the main crest of the outermost moraine (Fig. 7a). Additional data are needed to determine if this one age is an outlier or if this small crest could mark an earlier advance. The glaciers in the northern Chilean Lake District were extensive near this time at ~35 cal ka and ~32 cal ka [Denton *et al.*, 1999b]. The timing of the LGM advance is in good agreement with other Patagonian records ([Denton *et al.*, 1999b], [Douglass *et al.*, 2006], [Kaplan *et al.*, 2004], [Kaplan *et al.*, 2008] and [McCulloch *et al.*, 2005]). Figure 9 compares moraine boulder exposure ages calculated on a common basis from locations where detailed chronologies have been established for the LGM and/or Late-glacial period. The figure is restricted to cosmogenic-nuclide data with the main exception of northern Patagonia where few such data exist [Denton *et al.*, 1999b]. Across the northern Chilean Lake District, the most extensive LGM advance occurred at ~27 cal ka [22.5 ^{14}C ka], with the LGM maximum extent in the southern Chilean Lake District occurring later at ~17.9 cal ka [14.8 ^{14}C ka][Denton *et al.*, 1999b]. In the Strait of Magellan region, four recalculated ^{10}Be exposure ages obtained by McCulloch *et al.* [2005] and Kaplan *et al.* [2008] span a similar time period of 27 – 24 ka, with a mean of 25.5 ± 1.7 (Table 4). Thus overall, the timing of the LGM maximum extent appears to be broadly in phase throughout Patagonia.

The timing of the LGM maximum extent in the Lago Pueyrredón and nearby Lago Buenos Aires valleys exemplifies the benefit of obtaining multiple chronologies to better characterize the regional glacier response to climate. Specifically, the LGM maximum extent at Lago Buenos Aires (Fenix V moraines) is dated at ca. 4 ka later than the most extensive glacier extents in the Chilean Lake District ([Kaplan *et al.*, 2004] and [Douglass *et al.*, 2006]). The difference in the timing of the LGM maximum extent at Lago Buenos Aires (Fenix V: 23.8 ± 1.3) and Lago Pueyrredón is ~2 ka. If an earlier advance occurred at Lago Buenos Aires, the associated deposits have either been overridden during a later advance, or have been removed by subsequent meltwater erosion; thus the difference in the timing of the LGM maximum extent probably reflects a difference in the preservation of the glacial geologic record in the two valleys. Glasser and Jansson [2005] argue that fast-flowing outlet glaciers may have partially decoupled the icefield from climatically induced changes in ice thickness and extent. If so, the apparent difference in the glacier responses could be explained by different glaciological conditions unrelated to climate. Glasser and Jansson [2005] identify highly attenuated drumlins and flutes with mean length/width ratios of approximately 15:1 and 25:1 in the ice-scoured corridors of the Lago Pueyrredón/Cochrane and Lago Buenos Aires/General Carrera basins, respectively. These features indicate formation by fast-flowing ice ([Glasser and Jansson, 2005] and [Stokes and Clark, 2002]). The higher length/width ratio for landforms in the Lago Buenos Aires basin suggests that the flow of this outlet glacier may have been relatively faster [e.g., Stokes and Clark, 2002]. If there was a difference in ice-flux, the behaviour of the two glaciers should not be expected to be the same.

The initial stages of deglaciation began after ca. 18 – 17 ka. The timing of the final advance is similar to that of other Patagonian records (Fig. 9). For example, in the Chilean Lake District, the final advance is dated at 17.9 cal ka ([Denton *et al.*, 1999b], [Heusser *et al.*, 1999], [Lowell *et al.*, 1995] and [Moreno *et al.*, 1999]) and warming began at 17.5 cal ka ([Moreno and Leon, 2003] and [Massaferro *et al.*, 2009]). In the Strait of Magellan, the final advance occurred prior to ca. 17 cal ka based on minimum radiocarbon ages ([McCulloch and Davies, 2001] and [McCulloch *et al.*, 2005]). Recalculated ^{10}Be exposure ages for the final advance (Stage D) in the Strait of Magellan are 18.7 – 16.4 ka with a mean of 17.8 ± 2.5 ka. At Lago Buenos Aires, the recalculated ^{10}Be exposure ages for the Fenix I moraines range between 18.9 – 14.5 ka with a mean of 16.6 ± 1.2 ka. Bearing the uncertainties in mind, there appears to be broad agreement on the timing of the final advance prior to the onset of deglaciation throughout Patagonia at around 18 ka.

5.3. The evolution and diversion of paleolakes

The sequence and timing of deglaciation has implications for the development and drainage of a series of interconnected glacier-dammed lakes ([Bell, 2008] and [Turner *et al.*, 2005]). Based on the available ^{10}Be exposure ages ([Douglass *et al.*, 2005], [Douglass *et al.*, 2006] and [Kaplan *et al.*, 2004]) and radiocarbon ages [Turner *et al.*, 2005], we present a conceptual model of deglaciation for the region in Figure 10. The process of deglaciation in the Lago Pueyrredón valley was interrupted by a readvance or standstill that deposited the Lago Columna moraines (Fig. 10b). No direct ages for the Lago Columna moraines were obtained. However, based on our mapping and correlation with what is considered an equivalent ice limit in the Lago Pueyrredón valley (Fig. 2), we conclude that the age of the Lago Columna moraines must be younger than the ~18 – 17 ka Río Blanco moraine and that they were deposited at a time when ice occupied the Lago Pueyrredón valley and drainage of the pro-glacial lake was eastward via the Cañadón de Caracoles (Fig. 10b). Therefore, the Lago Columna moraines are likely older than the abandonment of this channel with a minimum age of around 16 ka. These data suggest that the Lago Columna moraines were deposited between ~18 – 16 ka.

In the Lago Buenos Aires valley, a readvance deposited the Menucos moraine to within a few kilometers of the Fenix I ice limit (Fig. 5b). The Menucos moraine occupies a similar morphostratigraphic position as the Lago Columna moraines, and they may be correlative. The ^{10}Be exposure ages for boulders on the Menucos moraine [Douglass *et al.*, 2006] give a recalculated age range from 16.5 to 13.6 ka, with a mean of 15.2 ± 1.3 ka. These data are consistent with radiocarbon ages obtained from the three carbonate-cemented concretions in varved lake sediments located stratigraphically between the Menucos and Fenix I moraines, which range from 15.0 ± 0.2 cal ka to 16.5 ± 0.5 cal ka ([Sylwan, 1989], [Kaplan *et al.*, 2004] and [Douglass *et al.*, 2006]). However, a note of caution is that the exact genesis of the carbonate concretions is uncertain and thus the level of accuracy of these ^{14}C ages.

Following deposition of the Menucos and Lago Columna moraines, the glaciers in both valleys withdrew by 80 – 140 kilometers, allowing Lago Pueyrredón to join Lago Buenos Aires as one united lake that drained eastward via the Río Deseado (Fig. 10c). If our assumptions are correct, the exposure age of ca. 16 ka for sample BC07-13, is an approximate minimum for the timing of this event. The age is consistent with radiocarbon ages of ~15.7 and 13.6 cal ka from the base of Cerro Ataud, which indicate that deglaciation had progressed enough to establish the lower united-paleolake by this time [Turner *et al.*, 2005]. The data are also consistent, within uncertainty, with the recalculated age for the Menucos moraines. Based on the surface exposure and radiocarbon ages presented, we estimate that widespread and rapid glacier retreat, including the formation and draining of the upper united-paleolake, occurred at around 16.5 – 15 ka as argued by Turner *et al.* [2005] on the basis of radiocarbon ages. Rapid glacier retreat is supported by glaciological modeling of the outlet glaciers [Hubbard *et al.,* 2005], and is consistent with rapid glacier retreat (ca. 80 km) reported in the Strait of Magellan near this time [McCulloch *et al.,* 2005]; the retreat was likely accelerated by an active calving margin. The proposed timing of retreat coincides with a time when temperatures in the Chilean Lake District had reached modern levels of warmth (ca. 15.6 - 15.3 cal ka; [Moreno *et al*., 1999]).

The history of the lower united-paleolake is less clear (Section 2.2). Further work is required to establish whether the lake reformed at some stage in the Holocene.

5.4. Wider implications

The overall structure of the LGM at Lago Pueyrredón bears similarities to other paleoclimate records in southern South America and in other parts of the world. The maximum ice extent occurred at ca. 27 – 25 ka and the glacial climate persisted over at least a 7 – 10 ka period as evident by the moraine ages, the last of which occurred at ca. 18 – 17 ka, before the onset of deglaciation. Glaciers reached their LGM maximum extent coincident with minimum Northern Hemisphere temperatures [e.g., GISP2; Alley, 2000], despite a peak in summer insolation intensity at middle latitudes in the Southern Hemisphere at this time [Berger, 1978] (Fig. 9). These data lend further support to the idea that global climates are broadly synchronized on orbital timescales [Denton *et al.,* 1999a].

During the glacial-interglacial transition, the rapid retreat of glaciers east of the North Patagonian Icefield stabilized or readvanced enough to maintain the glacier dam of the lower united-paleolake [Turner *et al.,* 2005]. Ice contact deposits associated with the lake have radiocarbon ages greater than ca. 12.8 cal ka [Turner *et al.,* 2005] that coincide with the timing of the Antarctic Cold Reversal at 14.8 – 12.7 ka [Blunier *et al.*, 1998]. Thus, in common with Antarctica, glaciers in central Patagonia may have responded to cooling in the Southern Hemisphere at the time. To simulate the glacier response during deglaciation, Hubbard *et al.* [2005] used an ice sheet model driven by a re-scaled Vostok temperature record; rapid retreat of the modeled glacier slows during the Antarctic Cold Reversal and retreats rapidly once more during the Younger Dryas. An interesting puzzle remains in that paleoecological records from the Península de Taitao located west of the North Patagonian Icefield (Fig. 1a), suggest that no significant cooling capable of reversing forest development occurred during either the Antarctic Cold Reversal or the Younger Dryas stadial at these latitudes ([Bennett *et al.*, 2000], [Hoganson and Ashworth, 1992], [Lumley and Switsur, 1993] and [Mercer, 1976]).

The regional glacier response during the time of the Northern Hemisphere's Younger Dryas stadial is equivocal. Mercer [1976] concludes that the Younger Dryas and Holocene advances remained within neoglacial limits, at least in the case of Glacier Témpano. Turner *et al.* [2005] suggest further that deglaciation may have actually begun during the Younger Dryas, which allowed the lower united lake to empty into the Pacific at the time. In contrast, Douglass *et al.* [2005] and Bell [2008] suggest that the glacier damming of the lower united lake may have persisted until as late as, or occurred at, ca. 6 ka (Fig. 10d)(Section 2.2). Paleoecological records from the Península de Taitao indicate the climate was somewhat cooler and drier during the

Younger Dryas [Massaferro and Brooks, 2002], but they do not support a significant climate reversal during this period ([Bennett *et al.,* 2000], [Hoganson and Ashworth, 1992] and [Lumley and Switsur, 1993]). While there remains uncertainty over the Younger Dryas response, collectively the available evidence in the region does not suggest a significant glacial advance at this time.

The glacial geologic and paleoecologic records in central Patagonia do not echo a distinct Antarctic or North Hemisphere (as observed in North Atlantic marine cores and Greenland ice cores) climate influence on millennial-scale climate variability during the last glacial-interglacial transition. In southernmost Patagonia and Tierra del Fuego (ca. 51 – 53º S), some glaciers responded in common with an Antarctic signal as evidenced by glaciers that readvanced during the Antarctic Cold Reversal and then retreating during the Younger Dryas stadial ([Fogwill and Kubik, 2005], [McCulloch *et al.,* 2005], [Moreno *et al.,* 2009] and [Sugden *et al.,* 2005]). However, the available proxy records in southern Patagonia do not present a clear late glacial scenario. A significant readvance of the South Patagonian Icefield is documented in the Lago Argentino area (~50º S) at around 13 ka, which formed the Puerto Bandera moraines ([Strelin and Denton, 2005] and [Strelin and Malagnino, 2000]). Recently, Ackert *et al.* [2008] inferred this event occurred at the end of, or after, the Younger Dryas stadial, based on ^{10}Be and ^{36}Cl exposure ages, coincident with peak temperatures in Antarctica (Fig. 9). In the Chilean Lake District of northwestern Patagonia (ca. 41 – 43º S), a more 'northern' climate signal has been inferred. Chironomid and pollen records indicate that the climate cooled with peaks at 14 cal ka and 13.5 cal ka [Massaferro *et al.,* 2009], and glaciers advanced at around 13.2 – 11.9 cal ka during the Huelmo/Mascardi Cold Reversal (11.4 – 10.2 ^{14}C ka), which is approximately coincident with the Younger Dryas stadial but began 550 years earlier

([Ariztegui *et al.*, 1997] and [Hajdas *et al.*, 2003]). The paleoecological records in central Patagonia (Península de Taitao) do not mirror either Antarctic or Northern Hemisphere climate influences during the glacial-interglacial transition. Thus, the central Patagonian records may reflect some blending of Antarctic and Northern Hemisphere influences as previously suggested for southernmost Patagonia [Sugden *et al.*, 2005].

6. Conclusions

Our mapping and cosmogenic ^{10}Be surface exposure ages allow us to draw the following conclusions:

1. The LGM maximum ice extent in the Lago Pueyrredón valley (~47.5º S) corresponds to the outermost limit of the Río Blanco moraine system; this maximum ice limit occurred by 27 – 25 ka, and is consistent with other Patagonian records in the Chilean Lake District (~42º S) and the Strait of Magellan (~53º S).

2. Slightly less-extensive Río Blanco ice limits are dated at 23 - 22 ka, 20 – 18 ka and ca. 18 – 17 ka. The overall timing of the LGM in the Lago Pueyrredón valley, with the maximum extent by 27 – 25 ka and deglaciation after ca. 18 – 17 ka, is broadly synchronous with other glacier chronologies in both hemispheres (e.g., [Clark *et al.*, 2009] and [Schaefer *et al.*, 2006]).
3. The new chronology indicates that the Río Blanco moraine system was not deposited during the Late-glacial to Holocene period as previously suggested [Wenzens, 2005].

4. Glaciers in both the Lago Buenos Aires and Lago Pueyrredón valleys retreated rapidly at around 16.5 – 15 ka. This retreat was associated with the evolution of a series of linked glacier-dammed lakes whose drainage shifted across the continental divide as a consequence of disintegrating ice.

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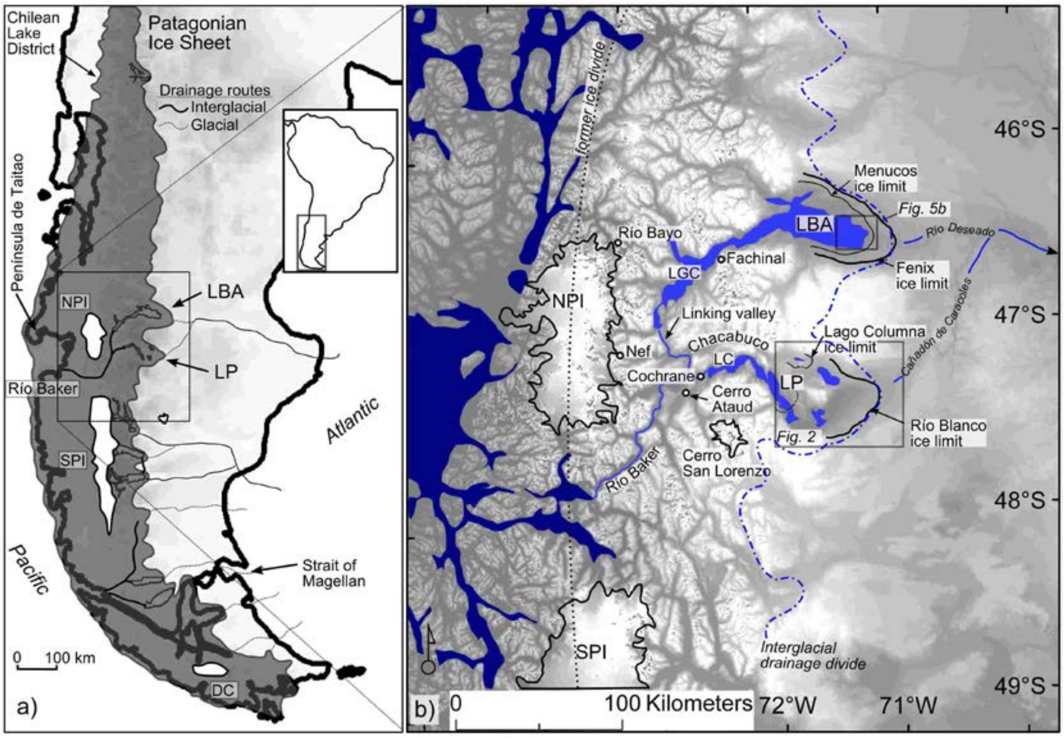

Figure 1. a) The general form of the former Patagonian Ice Sheet, showing locations discussed in the text including the location of the Lago Buenos Aires (LBA) and Lago Pueyrredón (LP) valleys, principal drainage routes and the present day North (NPI) and South (SPI) Patagonian Icefields and the Darwin Cordillera Icefield (DC).
b) The central Patagonian region including key locations mentioned in the text. Estimated position of the former ice divide and present drainage divide are shown to illustrate the different hydrologic conditions.

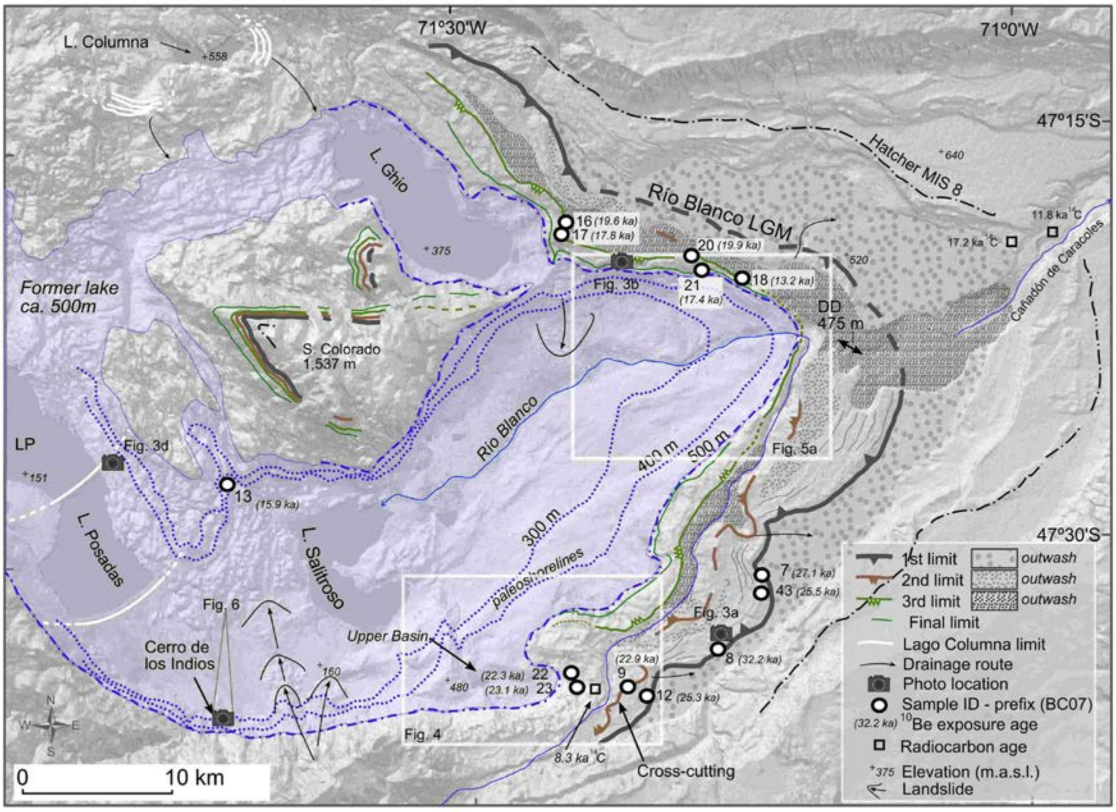

Figure 2. Geomorphic map of the Lago Pueyrredón (LP) valley with the principal ice limits, lake levels and sample sites identified in the text. The outermost moraine crest marking each ice limit is mapped. The outwash terraces are distinguished based on their associations with former ice limits; these slope gently eastward at ca. 0.5º. The radiocarbon ages are from Wenzens [2005] and Mercer [1982]. The base map is a SRTM 90m DEM.

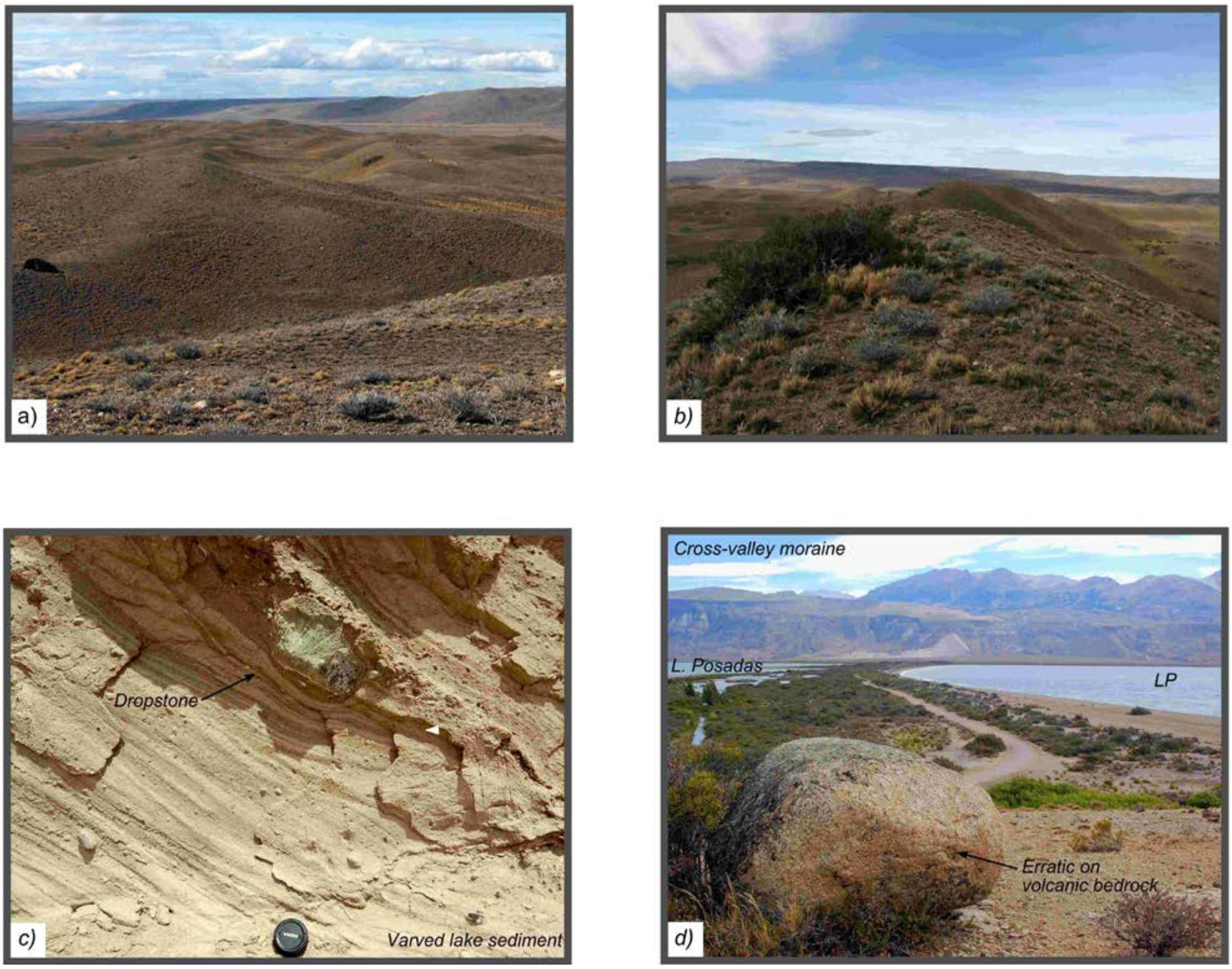

Figure 3. The locations of the photographs are shown in Figures 2 and 4.

a) The hummocky, continuous form of moraines associated with the 'First Limit.' Five to six moraine crests associated with the First Limit are preserved on the south side of the valley as shown in Figure 2.
b) The distinct sharp-crested continuous form typical of moraines marking the 'Third Limit.'
c) The photo shows deformed varved lake sediment containing dropstones. The deposit is located at ca. 570 meters in a former pro-glacial lake basin (Fig. 4). The section contains two sequences of till overlain by lake sediments, suggesting at least two advances occurred in the upper basin, each followed by the formation of a lake.
d) The cross-valley sub-lacustrine moraine separating Lago Posadas from Lago Pueyrredón (LP). The concentration of erratic boulders increases in close proximity to the causeway.

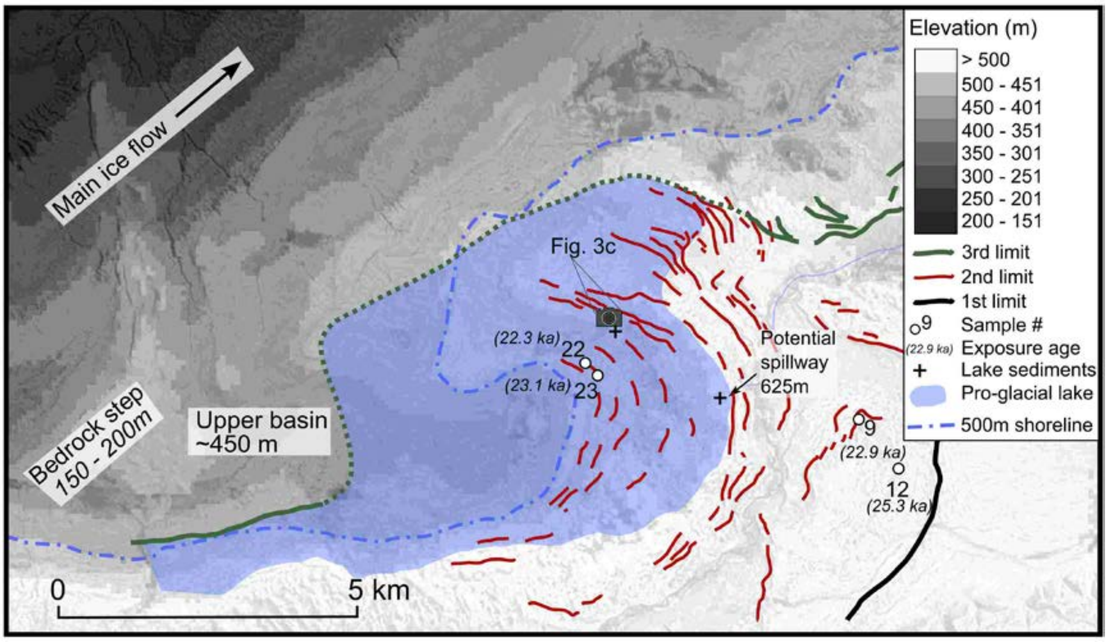

Figure 4. The southernmost part of the moraine system showing ice limits based on satellite imagery (location shown in Fig. 2). A bedrock step effectively isolates a basin from the main valley. The arcuate terminal moraines enclosing the basin are mantled by lacustrine sediments below 625 meters (See Fig. 3c). The inferred ice limit (dotted) and pro-glacial lake (at 625 meter elevation) associated with the Third Limit are shown. Samples BC07-22 and 23 would have been shielded by about 65 meters of water; their indistinguishable ages with sample BC07-9 suggests that the lake was relatively short-lived. We speculate that only the First and Second advances were voluminous enough to fully occupy the upper basin. Later advances were likely grounded to the west due to the relatively thin ice and a resultant lower ice flux.

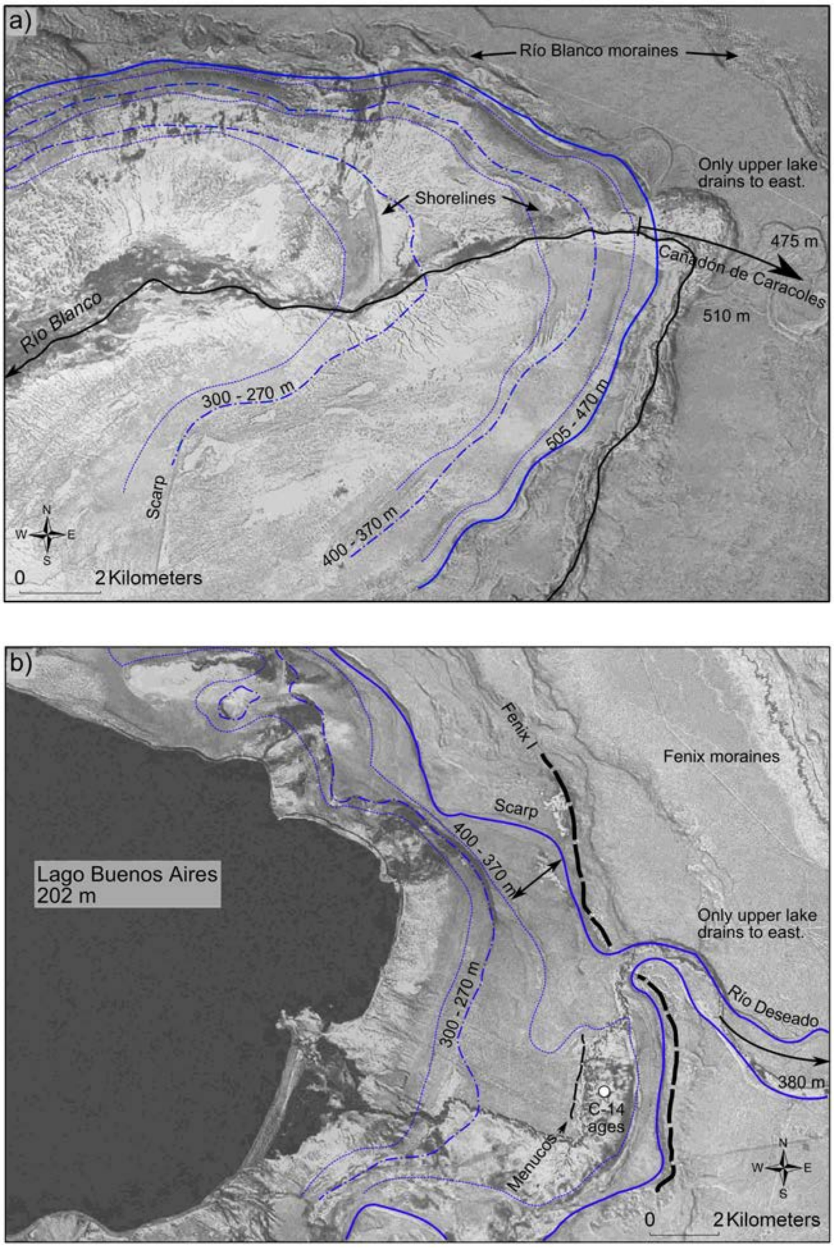

Figure 5

a) The principal paleoshoreline levels preserved in the Lago Pueyrredón valley (location shown in Figure 2). The upper Pueyrredón paleolake (505 – 470 m) drained to the east via the Cañadón de Caracoles (spillway ~475 m), the upper united-paleolake (400 – 370 m) drained to the east via the Río Deseado (spillway ~380 m) at Lago Buenos Aires (Fig. 5b) and the lower united-paleolake (300 – 270 m) drained to the west probably via the Río Baker which is assumed to have been glacier-dammed [cf. Bell, 2008].

b) The paleoshorelines representing the upper (400 – 370 m) and lower (300 – 270 m) united-paleolakes are also found east of Lago Buenos Aires (location shown in Figure 1b). Also shown are the location of the moraines and ^{14}C ages from carbonate concretions that are discussed in text.

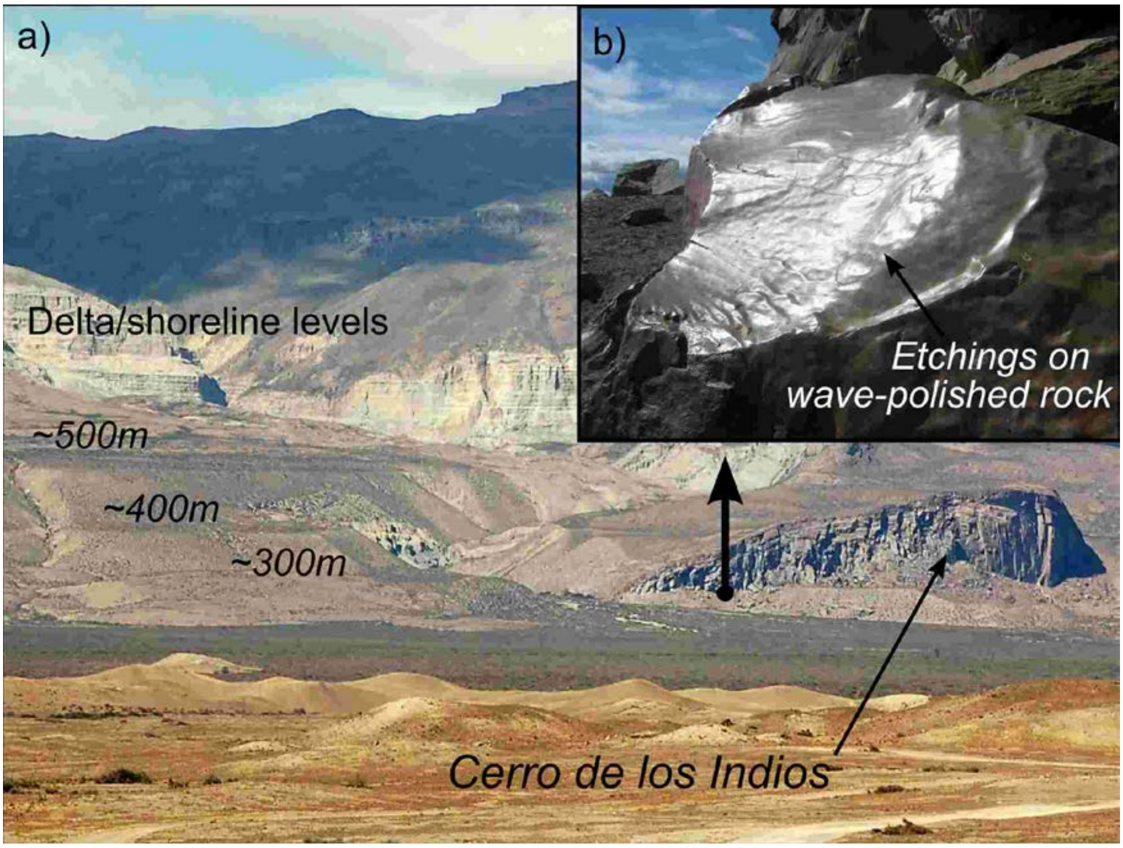

Figure 6.

a) Deltas associated with the three main paleoshoreline levels at Lago Pueyrredón. The location of the photograph is shown in Figure 2.
b) The smooth wave-polished boulders at the base of Cerro de los Indios (275m) are associated with the lowest united-paleolake. These were etched and painted by the Tehuelche Indians.

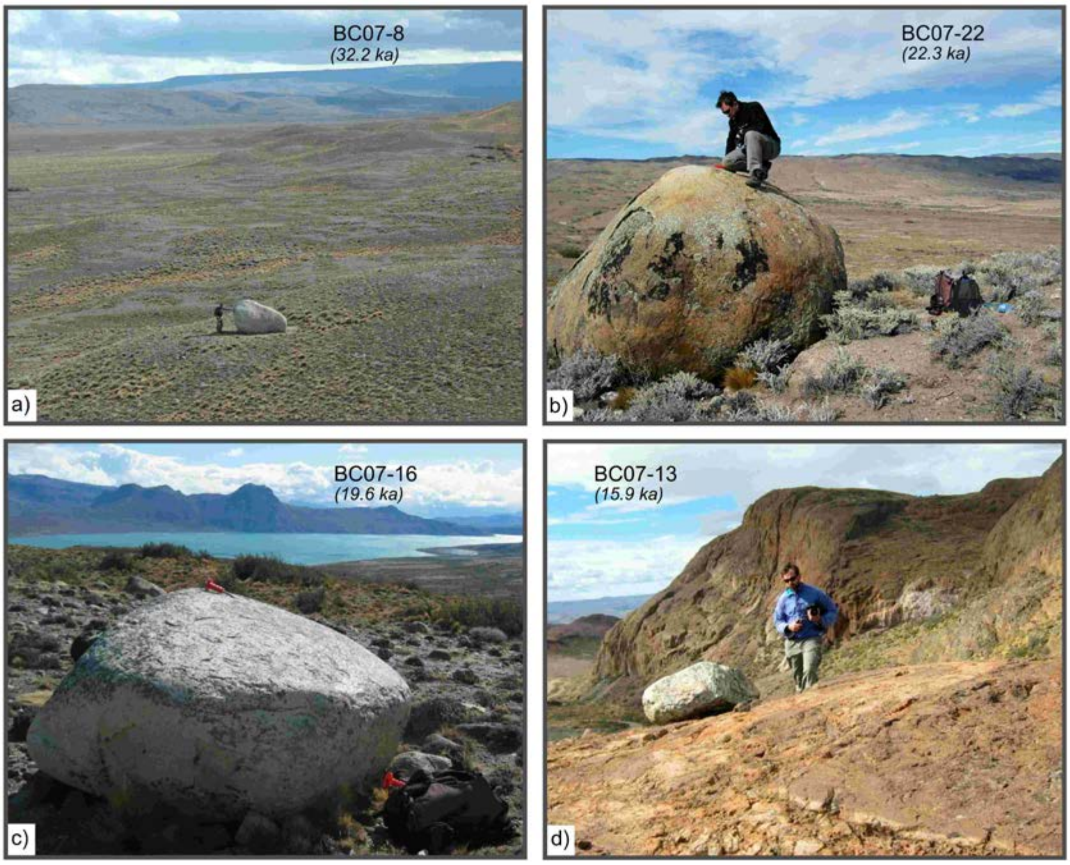

Figure 7.

Some of the boulders used for surface exposure dating (locations shown in Figure 2):

a) Large (height 1.9 m) boulder from the outermost position of the First Limit. This boulder is on a small crest exterior to the main crest; it could date an earlier advance however it is considered an outlier.
b) Large (height 2.5 m) boulder from the innermost moraine of the Second Limit showing evidence of spalling on the left side.
c) Fresh boulder (height 1.0 m) from the Third Limit near Lago Ghio.
d) Small boulder (height 0.75 m) resting on striated volcanic bedrock. This boulder was submerged in 100 – 150 meters of lake water following deposition.

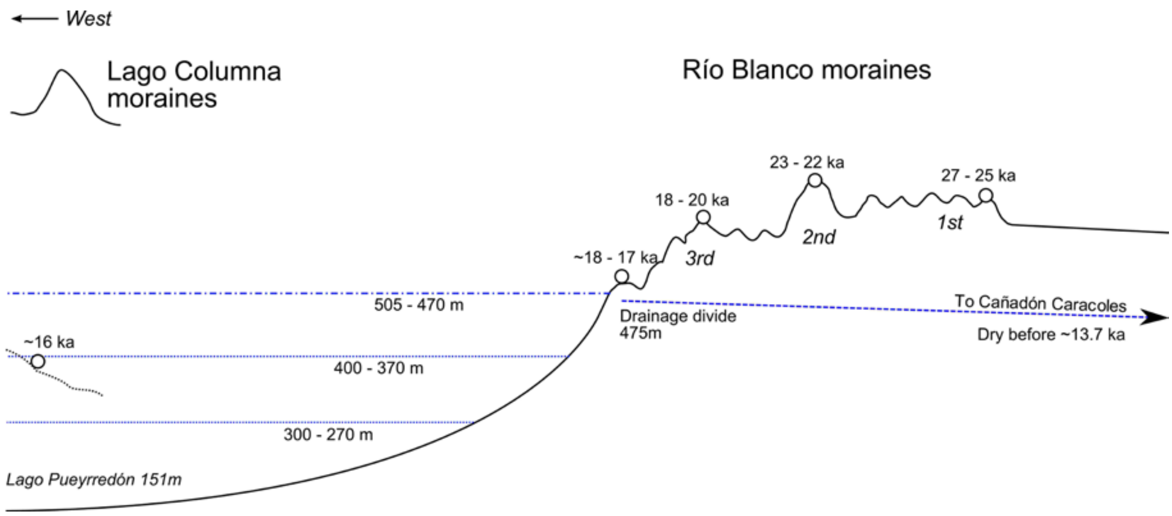

Figure 8. Approximate exposure ages and their geomorphic context (not to scale). The Cañadón de Caracoles dried after the lake level dropped to the 400m level; our exposure age of ca. 16 ka is consistent with the radiocarbon ages obtained by Mercer [1976, 1982] and Turner *et al.* [2005].

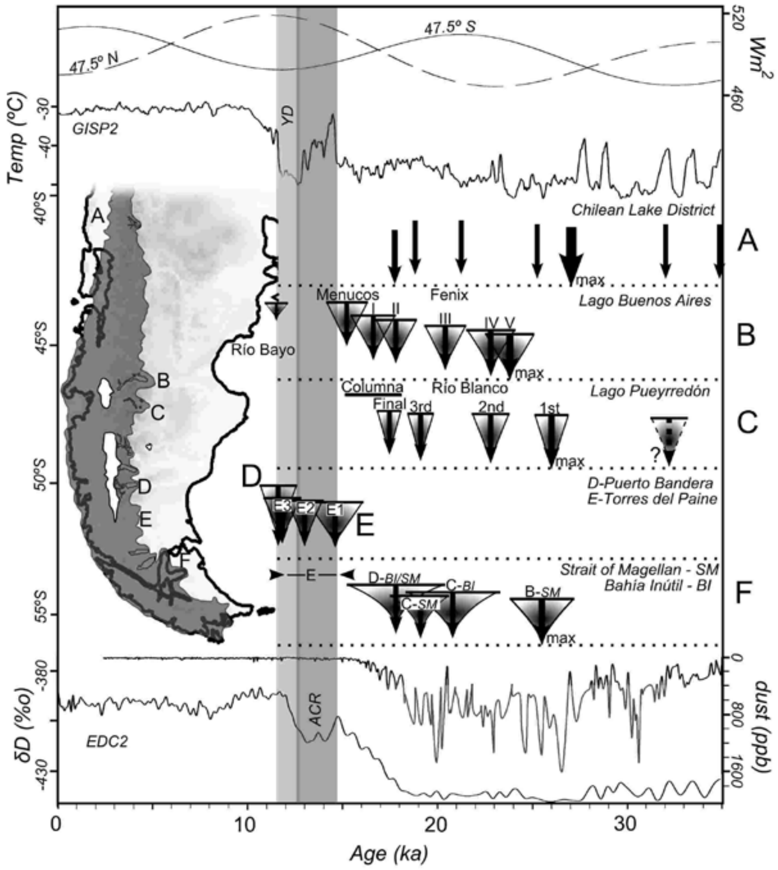

Figure 9. A comparison of available ^{10}Be exposure ages for moraines in Patagonia calculated on a common basis. The figure does not attempt to compare moraines with bracketing radiocarbon ages except where noted and thus only analytical uncertainties are presented. All ^{10}Be exposure ages were re-calculated with the CRONUS-Earth exposure age calculator (version 2.2; [Balco *et al.*, 2008]) according to the Dunai [2001] scaling model, and assumed no rock surface erosion. The mean age for each moraine is indicated by the arrow point, while the 1σ analytical uncertainty is indicated by the width of the triangle; the uncertainty is calculated as discussed in the text. The relative extent of each moraine-defined limit is shown decreasing from the LGM maximum limit which is denoted by the dotted horizontal line at each site. Data sources are: EDC2 deuterium and dust [EPICA *community members*, 2004]; Strait of Magellan and Bahía Inútil glacier advances B, C, D and E (note: stage E is based on bracketing radiocarbon ages) ([McCulloch *et al.,* 2005] and [Kaplan *et al.*, 2008]); Torres del Paine (note: E1 – is the 'outermost moraine', E2 – is 'within outermost moraine' and E3 – is the 'innermost moraine' as discussed by Moreno *et al.* [2009], and the data from both valleys have been averaged together ([Fogwill and Kubik, 2005] and [Moreno *et al.*, 2009]); the chronology for the Puerto Bandera moraines is by Ackert *et al.* [2008] (^{10}Be – Dmin) and ([Strelin and Malagnino, 2000] and [Strelin and Denton, 2005])(^{14}C – Dmax); Lago Pueyrredón (This study); Lago Buenos Aires ([Kaplan *et al.*, 2004] and [Douglass *et al.*, 2006]); Río Bayo is a minimum age [Glasser *et al.*, 2006]; radiocarbon ages from the Chilean Lake District [Denton *et al.*, 1999b]; GISP2 temperature [Alley, 2000]; Insolation [Berger, 1978].

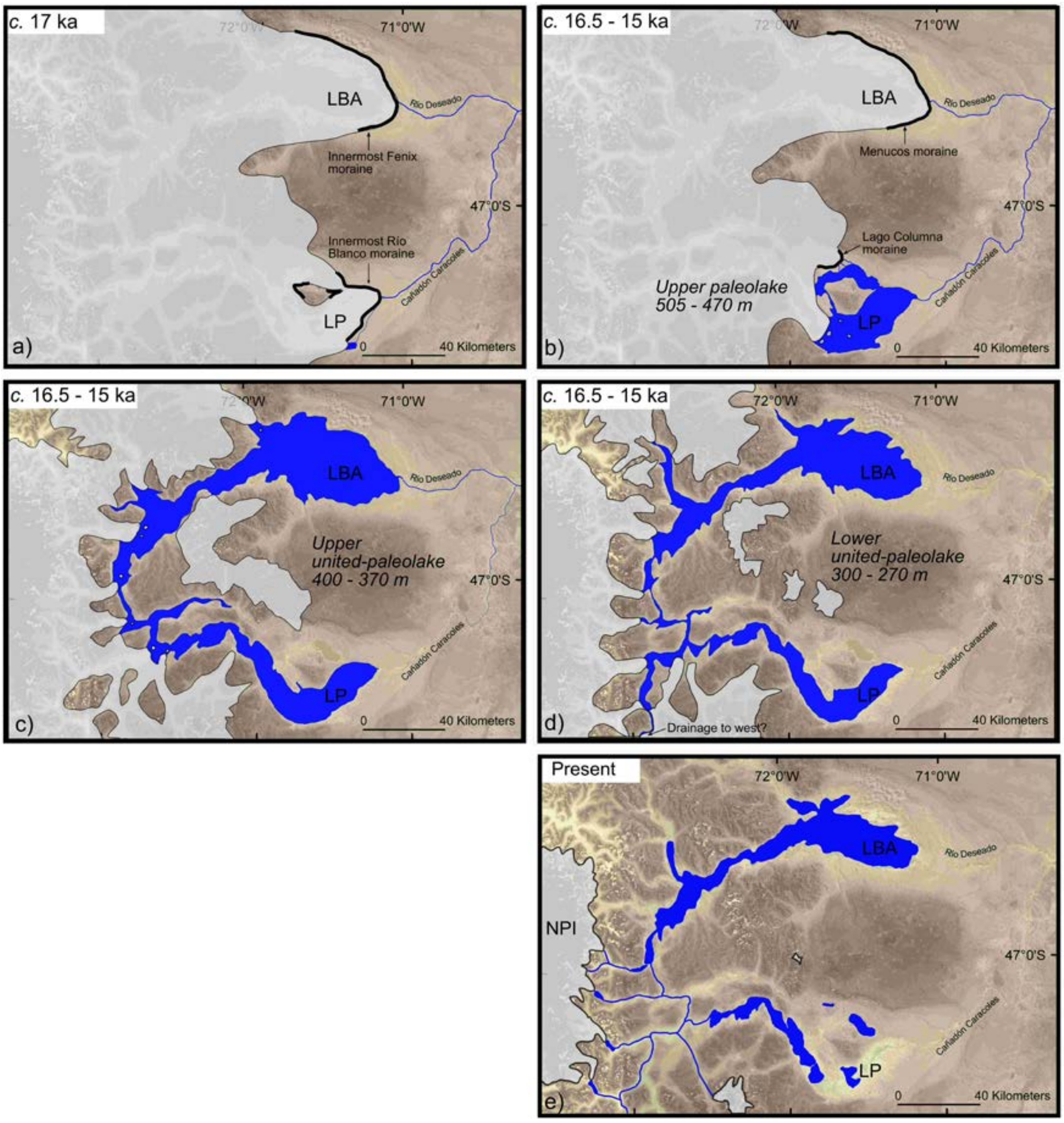

Figure 10. A conceptual model that approximates the pattern and timing of deglaciation in the region as discussed in the text. As such, it is not based on any known configuration of the outlet glaciers during the stated time intervals, except where noted and where ice margins are marked in bold and sources are cited. We make no attempt to distinguish the height of the ice sheet except where this was mapped in the Lago Pueyrredón (LP) valley, and thus we do not attempt to distinguish nunataks or other ice free areas in the interior.

a) ca. 17 ka: Glaciers fill the lake basins of Lago Buenos Aires (LBA) and Lago Pueyrredón and terminate at the innermost Fenix moraine ([Douglass *et al.*, 2006] and [Kaplan *et al.*, 2004]) and the innermost Río Blanco moraine, respectively.
b) 16.5 – 15 ka: Deglaciation is interrupted by a readvance that deposits the Menucos ([Douglass *et al.*, 2006] and [Kaplan *et al.*, 2004]) and Lago Columna moraines; the upper paleolake in the Lago Pueyrredón valley forms around this time.

c) 16.5 – 15 ka: The Lago Pueyrredón and Lago Buenos Aires outlet glaciers rapidly retreat to expose the valley linking the two lake basins, this allows the two lakes to join as one united lake (upper united-paleolake) that drains eastward via the Río Deseado.
d) 16.5 – 15 ka: The glaciers retreat sufficiently to allow westward drainage and establishment of the lower united-paleolake around this time.
e) The present day configuration was established when the Río Baker became ice free which allowed the lower united-paleolake to drain freely into the Pacific Ocean. Estimates on the timing range from 12.8 ka [Turner *et al.*, 2005] to after ca. 6.2 ka ([Bell, 2008] and [Douglass *et al.*, 2005]).

Table 1 Datasets.

Dataset		Spatial resolution (m)	Band combination	Source
LandSat TM	Imagery	30	7-red; 4-green; 2-blue	NASA's GeoCover: https://zulu.ssc.nasa.gov/mrsid/
ASTER	Imagery	30	3N-red; 2-green; 1-blue	Global Land Cover Facility: http://www.landcover.org/data/aster/[a]
SRTM	DEM	90	na	USGS EROS center: http://eros.usgs.gov/[b]

[a]USGS and Japan ASTER program (2003), Orthorectified image: ast14dmo_00308312005143541_20080130110256_7245, 08/31/2005.
[b]NASA's Shuttle Radar Topography Mission: see Farr et al. [2007].

Table 2: ^{10}Be boulder exposure ages from the Río Blanco moraine system based on five scaling models.

ε = 0 mm ka^{-1} Context	Sample I.D.	Lal (1991)/Stone (2000) ± 1σ (x10^{3}a)	Time-dependent L/S ± 1σ (x10^{3}a)	Lifton *et al.* (2005) ± 1σ (x10^{3}a)	Desilets *et al.* (2003, 2006) ± 1σ (x10^{3}a)	Dunai (2001) ± 1σ (x10^{3}a)	Dunai (2001) ε=1.4 mm ka^{-1} ± 1σ (x10^{3}a)	Analytical uncertainty ± 1σ (x10^{3}a)	External uncertainty ε = 0 mm ka^{-1} (Dunai, 2001) ± 1σ (x10^{3}a)
1st limit[a]	BC07-7	25.7	25.8	26.3	27.0	27.1	*28.0*	0.9	3.3
	BC07-8*	30.6	30.6	31.0	32.0	32.2	*33.4*	1.0	3.9
	BC07-12	24.0	24.2	24.6	25.3	25.3	*26.1*	0.8	3.1
	BC07-43	24.1	24.3	24.7	25.4	25.5	*26.2*	0.8	3.1
	Arithmetic mean:	*24.6 ± 0.9*	*24.8 ± 0.9*	*25.2 ± 0.9*	*25.9 ± 0.9*	*26.0 ± 0.9*	*26.8 ± 0.9*		
2nd limit	BC07-9	21.7	21.8	22.2	22.8	22.9	*23.5*	0.7	2.8
	BC07-22	21.0	21.2	21.7	22.2	22.3	*22.9*	0.8	2.8
	BC07-23	21.8	22.0	22.4	23.0	23.1	*23.7*	1.3	3.0
	Arithmetic mean:	*21.5 ± 0.9*	*21.7 ± 1.0*	*22.1 ± 0.9*	*22.7 ± 1.0*	*22.8 ± 1.0*	*23.4 ± 1.0*		
3rd limit	BC07-16	18.5	18.7	19.1	19.6	19.6	*20.1*	0.6	2.4
	BC07-17	16.8	17.0	17.4	17.8	17.8	*18.2*	0.8	2.3
	BC07-18*	12.4	12.6	12.9	13.2	13.2	*13.4*	0.6	1.7
	BC07-20	18.8	19.0	19.4	19.9	19.9	*20.4*	0.6	2.4
	Arithmetic mean:	*18.0 ± 0.7*	*18.2 ± 0.7*	*18.6 ± 0.7*	*19.1 ± 0.7*	*19.1 ± 0.7*	*19.5 ± 0.7*		
Final limit	BC07-21	16.4	16.6	17.0	17.4	17.4	*17.8*	0.7	2.2
Lake draining	BC07-13	14.9	15.1	15.5	15.8	15.9	*16.1*	0.8	2.0

[a]Published by Hein et al. (2009).
* Sample excluded from calculation of arithmetic mean age.
Exposure ages calculated with the CRONUS-Earth web-based calculator (v.2.2; http://hess.ess.washington.edu/math/; Balco et al., 2008).
Exposure ages calculated assuming a rock density of 2.7 g cm^{-3} and no correction for erosion is applied except where noted.
Shielding negligible for all samples except BC07-13 (0.996).

Table 3: ^{10}Be data for the Río Blanco moraines at Lago Pueyrredón.

Sample I.D.	AMS I.D.	Latitude (dd)	Longitude (dd)	Altitude (m asl)	Boulder height (cm)	Thickness (cm)	Carrier mass (1000 ppm) (mg)	Quartz mass (g)	^{10}Be concentration[a] $\pm 1\sigma$[b] (10^5 atom g^{-1})
<u>Río Blanco</u>									
1st limit[c]									
BC07-7	b2050	-47.5186	-71.2361	564	225	2.5	0.2408	20.9284	2.00 ± 0.07
BC07-8	b2043	-47.5564	-71.2629	587	190	1.5	0.2465	27.7948	2.45 ± 0.08
BC07-12	b2036	-47.5840	-71.3548	665	310	1.5	0.2462	24.1815	2.07 ± 0.07
BC07-43	b2055	-47.5079	-71.2451	581	80	1.5	0.2524	25.7394	1.93 ± 0.06
2nd limit									
BC07-9	b2079	-47.5796	-71.3582	709	95	3	0.2461	27.0887	1.91 ± 0.06
BC07-22	b2090	-47.5726	-71.4171	544	250	1.5	0.2353	31.7051	1.63 ± 0.06
BC07-23	b2951	-47.5758	-71.4126	541	130	3.5	0.2463	8.5658	1.66 ± 0.10
3rd limit									
BC07-16	b2953	-47.3041	-71.4029	620	100	1	0.246	26.0982	1.53 ± 0.05
BC07-17	b1766	-47.3087	-71.4050	614	100	2	0.247	27.8613	1.37 ± 0.06
BC07-18	b1770	-47.3380	-71.2490	570	140	2	0.2463	27.9337	0.97 ± 0.05
BC07-20	b2954	-47.3259	-71.2938	565	75	2	0.2456	31.3211	1.47 ± 0.04
Final Advance									
BC07-21	b2051	-47.3361	-71.2830	509	95	2	0.246	28.527	1.23 ± 0.05
<u>Lake draining</u>									
boulder on bedrock									
BC07-13*	b2042	-47.4525	-71.7120	352	75	1.5	0.2461	30.9626	0.967 ± 0.051

* No correction for topographic shielding is applied to the data (horizons <~3º), except for sample BC07-13: the correction is 0.996.

[a]Measurements normalised to NIST SRM-4325 Be standard with revised ^{10}Be/^{9}Be ratio of 2.79 x 10^{-11} and half-life 1.36 Ma (Nishiizumi et al., 2007).

Concentrations corrected for process blanks, which ranged from 115 ± 18 x10^3 atoms ^{10}Be to 290 ± 40 x10^3 atoms ^{10}Be (<10% of total ^{10}Be atoms in the sample; ^{10}Be/

[b]Uncertainties include propogated sample and blank ^{10}Be/^{9}Be analytical uncertainties and a 2% carrier addition uncertainty.

[c]Published by Hein et al. (2009). All AMS measurements made at the Scottish Universities Environmental Research Centre (S.U.E.R.C.)

Table 4: Recalculated ^{10}Be exposure ages from Lago Buenos Aires and the Strait of Magellan

Context	Sample I.D.	Lal (1991)/Stone (2000) ± 1σ ($\times10^3$a)	Time-dependent L/S ± 1σ ($\times10^3$a)	Lifton et al. (2005) ± 1σ ($\times10^3$a)	Desilets et al. (2003, 2006) ± 1σ ($\times10^3$a)	Dunai (2001) ± 1σ ($\times10^3$a)	Dunai (2001) *ε=1.4 mm ka^{-1}* ± 1σ ($\times10^3$a)	Analytical uncertainty ± 1σ ($\times10^3$a)	External uncertainty ε = 0 mm ka^{-1} (Dunai, 2001) ± 1σ ($\times10^3$a)
ε = 0 mm ka^{-1}									
Lago Buenos Aires[a]									
Fenix V	LBA-01-60[a]	24.4	24.5	25.1	25.7	25.8	26.6	1.0	3.2
	LBA-02-20[b]	23.2	23.3	23.9	24.5	24.5	25.2	1.6	3.3
	LBA-98-135[a]	22.3	22.4	23.0	23.5	23.6	24.2	0.8	2.9
	LBA-01-22[a]	22.2	22.3	22.9	23.4	23.5	24.1	0.8	2.9
	LBA-02-21[b]	22.0	22.2	22.7	23.2	23.3	23.9	2.3	3.7
	LBA-02-24[b]	21.0	21.1	21.7	22.1	22.2	22.8	0.9	2.8
	LBA-02-22[b]*	17.9	18.1	18.6	19.0	19.0	19.4	1.3	2.6
	LBA-02-23[b]*	16.5	16.6	17.1	17.5	17.5	17.8	0.9	2.3
	Arithmetic mean:	*22.5 ± 1.3*	*22.6 ± 1.3*	*23.2 ± 1.3*	*23.7 ± 1.3*	*23.8 ± 1.3*	*24.5 ± 1.3*		
Fenix I	LBA-01-06[a]	17.8	18	18.5	18.9	18.9	19.3	1.4	2.7
	LBA-02-11[b]	16.6	16.7	17.2	17.6	17.6	17.9	1.3	2.5
	LBA-03-23[b]	16.5	16.7	17	17.4	17.4	17.7	1.1	2.3
	LBA-01-04[b]	16.4	16.6	17.1	17.4	17.4	17.8	0.9	2.3
	LBA-02-10[b]	15.6	15.8	16.3	16.6	16.6	16.9	0.9	2.2
	LBA-01-05[a]	15.2	15.4	15.8	16.1	16.1	16.5	0.5	2.0
	LBA-02-13[b]	14.6	14.8	15.3	15.6	15.6	15.9	2.0	2.8
	LBA-02-12[b]	14.3	14.5	14.9	15.2	15.2	15.4	1.6	2.5
	LBA-03-19[b]	13.8	13.9	14.3	14.6	14.5	14.8	1.0	2.0
	LBA-03-21[b]*	10	10.2	10.5	10.7	10.6	10.8	1.0	1.6
	Arithmetic mean:	*15.6 ± 1.2*	*15.8 ± 1.2*	*16.3 ± 1.2*	*16.6 ± 1.2*	*16.6 ± 1.2*	*16.9 ± 1.2*		
Menucos	LBA-03-11[b]	15.6	15.8	16.2	16.5	16.5	16.8	3.0	3.7
	LBA-04-12[b]	14.9	15.1	15.5	15.8	15.8	16.0	1.0	2.2
	LBA-03-13[b]	14.8	15	15.4	15.7	15.7	15.9	1.0	2.1
	LBA-04-14[b]	14.8	15	15.3	15.7	15.6	15.9	0.8	2.0
	LBA-03-10[b]	13.3	13.4	13.8	14.1	14.0	14.3	1.2	2.1
	LBA-03-17[b]	12.9	13	13.4	13.7	13.6	13.8	0.9	1.8
	LBA-03-14[b]*	10.6	10.8	11.1	11.3	11.2	11.4	0.7	1.5
	LBA-03-15[b]*	10.7	10.9	11.2	11.4	11.4	11.5	0.9	1.6
	LBA-04-13[b]*	10.8	11	11.2	11.5	11.4	11.6	0.6	1.5
	Arithmetic mean:	*14.4 ± 1.3*	*14.6 ± 1.3*	*14.9 ± 1.3*	*15.3 ± 1.3*	*15.2 ± 1.3*	*15.5 ± 1.3*		
Strait of Magellan									
Stage B	BGG-C1[c]	25.3	25.8	25.6	26.6	26.7	27.4	1.9	3.8
	BGG-C2[c]	24.2	24.7	24.6	25.5	25.6	26.3	2.4	4.0
	BGG-C3[d]	23.9	24.5	24.3	25.3	25.4	26.0	1.0	3.2
	BGG-C4[c]	22.8	23.3	23.1	24.0	24.1	24.7	1.4	3.2
	Arithmetic mean:	*24.0 ± 1.7*	*24.6 ± 1.7*	*24.4 ± 1.7*	*25.4 ± 1.7*	*25.5 ± 1.7*	*26.1 ± 1.7*		
Stage D	SM-02-21[d]	17.7	18.1	18.0	18.7	18.7	19.2	5.2	5.9
	BI-C3[c]	17.2	17.6	17.5	18.1	18.2	18.6	1.9	2.9
	BI-C2[c]	17.1	17.5	17.4	18.0	18.1	18.5	2.6	3.5
	BI-C1[c]	16.4	16.7	16.7	17.3	17.3	17.7	1.4	2.5
	SM-02-7[d]	17.0	17.3	17.3	17.8	17.9	18.3	1.1	2.4
	SM-02-8[d]	17.2	17.6	17.5	18.1	18.2	18.5	4.1	4.8
	SM-02-23[d]	15.5	15.8	15.8	16.4	16.4	16.7	1.5	2.5
	Arithmetic mean:	*16.9 ± 2.6*	*17.2 ± 2.6*	*17.2 ± 2.6*	*17.8 ± 2.6*	*17.8 ± 2.6*	*18.2 ± 2.6*		

[a] Kaplan et al. (2004) [b] Douglass et al. (2006)
[c] McCulloch et al. (2005) [d] Kaplan et al. (2008)

* Sample excluded from calculation of arithmetic mean age following original publications.

All exposure ages calculated with the CRONUS-Earth web-based calculator (v.2.2; http://hess.ess.washington.edu/math/; Balco et al., 2008).

Rock density and sample shielding as per original publications.